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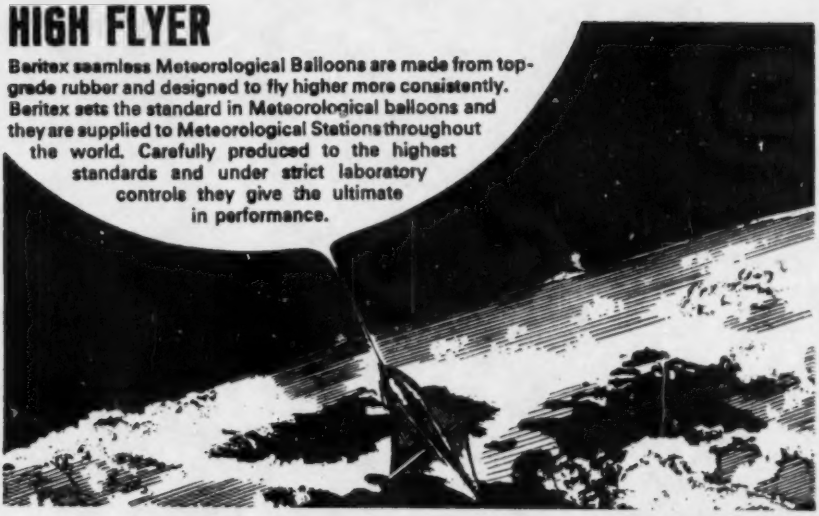

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ON CERTAIN LARGE-SCALE SYNOPTIC FEATURES OF THE SUMMER OVER THE BRITISH ISLES AND AUTUMN MEAN TEMPERATURE IN CENTRAL ENGLAND AND AUTUMN RAINFALL OVER ENGLAND AND WALES

By R. MURRAY

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To surmount some of these drawbacks of synoptic investigations it is helpful to be able to represent the dominant synoptic characteristics of a season by synoptic parameters. Murray and Lewis[2] have already presented synoptic indices, based on the Lamb[3] catalogue of daily synoptic types over the British Isles. These indices are quite objective once the daily synoptic types have been selected; any uncertainty in the synoptic-type selections is not in general of much importance when seasonal quintile values of the indices are considered, as is done in this note. Broadly speaking, the *P*-index over a period measures the overall difference in frequency of days with progressive (mainly westerly) synoptic types and non-progressive (mainly easterly or meridional) synoptic types over the British Isles; *P* in quintile 5 refers to

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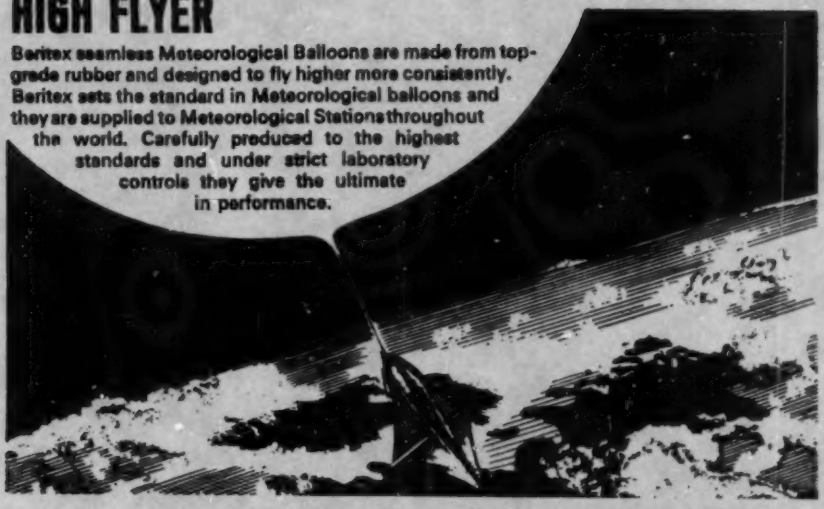

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the most progressive class. The *S*-index is a measure of the bias towards southerly or northerly synoptic types; *S* in quintile 5 refers to the most southerly (or least northerly) class. The *C*-index is effectively a measure of the difference in frequency of cyclonic and anticyclonic days near the British Isles; *C* in quintile 5 refers to the most cyclonic (or least anticyclonic) class. For precise details of the method whereby the indices are computed, the earlier paper[2] should be consulted. Quintile boundaries of the *P*, *S* and *C* indices for the summer and the July to August periods are given in an appendix to this paper.

Tentative notions concerning summer synoptic types and autumn temperature. The writer had noticed that the summers in the 1960's (except 1967) appeared to be much more northerly in character than usual for the season and that the mean temperatures in central England in the autumns were generally near or below the seasonal average (i.e. they were not in quintiles 4 or 5). It was guessed that the northerliness or southerliness of the summer might be a predictor of autumn temperature. A quick run through the series of *S*-indices for the summers since 1873 showed that there were 36 summers of quintile 1 or 2 (S_1 or S_2) in this index (i.e. northerly bias) and the following autumn mean temperatures in central England were below average (T_1 or T_2) on 22 occasions and above average (T_4 or T_5) on 10 occasions with the rest near average. There was evidently a tendency to have colder autumns after northerly summers, but the association was not pronounced enough for predictive usefulness. In the case of southerly summers the bias to warmer autumns was very slight. A similar check of the non-progressive or blocked summers (P_1 or P_2) showed that autumn temperatures in central England following these particular summers were twice as frequently below average as above average. On the other hand, progressive summers were almost as likely to be followed by cold autumns as by warm autumns. The cyclonic character of the summer seemed to be related to the autumn temperatures only in the extreme cases. Very cyclonic or C_5-summers were followed by mean autumn temperatures which were below, near and above average on 10, 4 and 4 occasions (none of the above-average cases was quintile 5). Very anticyclonic or C_1-summers showed a weak tendency to be followed in the autumn by mean temperatures above rather than below average. These considerations of the individual indices suggest that northerly and/or cyclonic and/or blocked summers on the one hand, and southerly and/or anticyclonic and/or progressive summers on the other hand might be precursors of quite different weather in the following autumns. These conjectures are examined further in the following sections.

Blocked, northerly, cyclonic summers and autumn weather. In this note, a blocked/northerly summer is taken as one in which the summer *P*-index is either quintile 1 (P_1) or quintile 2 (P_2) *and* the summer *S*-index is either quintile 1 (S_1) or quintile 2 (S_2). Thus a blocked/northerly summer might be $P_1\ S_1$ or $P_1\ S_2$ or $P_2\ S_1$ or $P_2\ S_2$ (for brevity, $P_{12}\ S_{12}$). Similarly blocked/cyclonic summers are those characterized by $P_1\ C_5$ or $P_1\ C_4$ or $P_2\ C_5$ or $P_2\ C_4$; and cyclonic/northerly summers are those characterized by $C_5\ S_1$ or $C_5\ S_2$ or $C_4\ S_1$ or $C_4\ S_2$. By means of the seasonal values of the *P*, *S* and *C* indices in quintile form, summers with these characteristics can be objectively selected from the historical record (1873 to 1967). There were

only 8 summers (namely 1910, 1912, 1948, 1957, 1960, 1963, 1965 and 1966) which were simultaneously blocked, cyclonic and northerly; on the other hand, 30 summers were characterized by at least one of the types $P_{12}\ S_{12}$, $P_{12}\ C_{45}$ and $C_{45}\ S_{12}$. In Table I the various categories of summers are related to temperature and rainfall data in the following autumns.

TABLE I—FREQUENCIES OF AUTUMN TEMPERATURE (QUINTILES) IN CENTRAL ENGLAND AND AUTUMN RAINFALL (TERCILES) OVER ENGLAND AND WALES FOLLOWING BLOCKED, NORTHERLY, CYCLONIC SUMMERS AS SPECIFIED BY SEASONAL *P*, *S* AND *C* INDICES (QUINTILES)

Type of summer	Autumn Central England temperature					Autumn England and Wales rainfall			No.
	1	2	3	4	5	1	2	3	
(a) blocked/northerly ($P_{12}S_{12}$)	4	4	2	0	0	3	5	2	10
(b) blocked/cyclonic ($P_{12}C_{45}$)	5	6	3	1	0	4	8	3	15
(c) cyclonic/northerly ($C_{45}S_{12}$)	7	5	3	5	1	7	9	5	21
(d) simultaneously *(a)* and *(b)* and *(c)*	3	3	2	0	0	2	5	1	8
(e) blocked/northerly and/or blocked/cyclonic	6	7	3	1	0	5	8	4	17
(f) at least one criterion *(a)*, *(b)* or *(c)* satisfied	10	9	4	6	1	11	11	8	30

Note : *(a)* blocked/northerly = P_1S_1 or P_1S_2 or P_2S_1 or P_2S_2 (i.e. $P_{12}S_{12}$);
(b) blocked/cyclonic = P_1C_5 or P_1C_4 or P_2C_5 or P_2C_4 (i.e. $P_{12}C_{45}$);
(c) cyclonic/northerly = C_5S_1 or C_5S_2 or C_4S_1 or C_4S_2 (i.e. $C_{45}S_{12}$);
Quintile boundaries of temperature and tercile boundaries of rainfall in the autumn are listed in reference 4.

The frequency distributions of autumn temperatures shown in Table I emphasize the strong bias to below average temperatures in central England following the various groups of summers. The least satisfactory prediction of cold autumns arises from the use of the $C_{45}\ S_{12}$ criterion (item *(c)*). If the stringent criteria *(d)* are demanded, then none of the following autumns is warm, but predictions would then only be possible on less than 10 per cent of occasions. Very satisfactory predictions result from the criteria that the summer be classified as $P_{12}\ S_{12}$ and/or $P_{12}\ C_{45}$ (item *(e)*); of 17 occasions only one autumn was in the warmer class (quintile 4). Relaxation of the summer requirements still further gives the distribution shown as *(f)* in Table I; this enables predictions to be made on 30 per cent of occasions, but the accuracy falls off somewhat.

As regards autumn rainfall over England and Wales the frequency distributions shown in Table I are not striking, but all except *(c)* and *(f)* suggest that average rainfall is most likely. Criteria *(c)* and *(f)* involve cases of cyclonic/northerly summers and these appear to be followed by a random distribution of rainfall terciles in the autumn.

In many summers the character of the whole season is dominated by the synoptic character of July and August, and it seemed desirable to see whether selections made on the basis of the *P*, *S* and *C* indices for the two-month period July to August would result in autumn temperature and rainfall frequency distributions similar to those shown in Table I. In general this

hypothesis was confirmed, but an interesting result emerged in the $C_{45}\ S_{12}$ case. As shown in Table I (item (*c*)), there were 6 warm autumns which followed the 21 $C_{45}\ S_{12}$ summers. When the basis of selection was limited to years with cyclonic/northerly (or $C_{45}\ S_{12}$) July to August periods (i.e. no account taken of the synoptic character of June) there resulted 17 selections. These 17 years were followed by 6, 7, 3, 1 and 0 occasions of temperature quintiles 1, 2, 3, 4 and 5 respectively, and by 6, 9 and 2 occasions of rainfall terciles respectively. Thus only one warm autumn followed the 17 years with July to August characterized by the synoptic index $C_{45}\ S_{12}$. In view of this result and the information contained in Table I, it is clear that years selected on the basis of $P_{12}\ S_{12}$ and/or $P_{12}\ C_{45}$ summers and/or $C_{45}\ S_{12}$ July plus August must tend to be followed by colder than usual autumns in central England. Some years may of course be selected by more than one of these synoptic parameters. The summarized results are given in Table II.

TABLE II—AUTUMN MEAN TEMPERATURES (QUINTILES) IN CENTRAL ENGLAND AND AUTUMN RAINFALL (TERCILES) OVER ENGLAND AND WALES AFTER BLOCKED/NORTHERLY (I.E. $P_{12}\ S_{12}$) SUMMERS AND/OR BLOCKED/CYCLONIC (I.E. $P_{12}\ C_{45}$) SUMMERS AND/OR CYCLONIC/NORTHERLY (I.E. $C_{45}\ S_{12}$) JULY PLUS AUGUST. SELECTION ACCORDING TO A SPECIFIC SYNOPTIC CHARACTERISTIC IS SHOWN BY AN ASTERISK

Year	Summer $P_{12}S_{12}$	Summer $P_{12}C_{45}$	July + August $C_{45}S_{12}$	Autumn temperature (quintiles)					Autumn rainfall (terciles)		
				1	2	3	4	5	1	2	3
1879		*		1					1		
1885	*			1							3
1888		*			2					2	
1889		*	*		2				1		
1907			*				4			2	
1910	*	*	*	1						2	
1912	*	*	*	1					1		
1915			*	1					1		
1917		*			2					2	
1922			*	1					1		
1927		*				3					3
1931			*		2					2	
1936			*		2					2	
1944		*		1							3
1948	*	*	*			3			1		
1952			*	1						2	
1954			*			3					3
1955	*				2				1		
1956			*		2				1		
1957	*	*	*		2					2	
1958		*					4			2	
1960	*	*	*		2						3
1963	*	*	*			3				2	
1965	*	*	*	1						2	
1966	*	*	*		2					2	
	Frequency distribution			9	10	4	2	0	8	12	5

The frequency distribution of autumn temperatures given in Table II shows a striking bias to quintiles one and two, with only 2 out of 25 occasions being above average; in fact the frequency distribution is highly significant statistically (at 0·5 per cent level by a chi-square test). The autumn rainfall pattern is not, however, statistically significant, even though it appears that the most likely rainfall is average.

The pre-conditions given by the combination of synoptic parameters in Table I (particularly items (*c*), (*b*), (*d*) and (*e*)) and Table II suggest strongly that cold autumns will very probably follow, but this does not mean that all the autumn months are likely to be cold. In Table III are presented the frequency distributions of temperature quintiles in the autumn months for the years examined in Tables I and II.

TABLE III—FREQUENCY OF MONTHLY MEAN TEMPERATURE IN QUINTILES FOR SEPTEMBER, OCTOBER AND NOVEMBER FOLLOWING SPECIFIED SYNOPTIC TYPES IN PRECEDING SUMMER

Synoptic type	No.	September					October					November				
		1	2	3	4	5	1	2	3	4	5	1	2	3	4	5
(i) $P_{12}S_{12}$ summer	10	4	3	2	1	0	2	1	4	2	1	3	1	3	2	1
(ii) $P_{12}C_{45}$ summer	15	4	7	2	1	1	4	2	5	3	1	5	4	3	3	2
(iii) $C_{45}S_{12}$ July+August	17	7	4	3	2	1	4	4	5	2	2	5	4	3	3	2
(iv) at least one of (i), (ii) or (iii)	25	9	7	3	4	2	8	6	6	3	2	6	6	6	3	4

It is clear from Table III that the bias to cold is most pronounced in September; there is a random distribution of temperature quintiles in November.

Progressive, southerly, anticyclonic summers and autumn weather. The types of summer discussed briefly in this section are synoptically quite different from the summers analysed in the preceding section. It might therefore be expected that the autumn temperature and rainfall following progressive, southerly, anticyclonic summers would be different from the autumn weather following blocked, northerly, cyclonic summers.

Progressive/anticyclonic summers, defined strictly as P_{45} C_{12}, occurred on only 8 occasions; it was necessary to allow some relaxation of P or C to quintile 3 in order to obtain a larger number of occasions. Redefining progressive/anticyclonic summers as those of the type P_{345} C_{12} or P_{45} C_{123} increased the number of occasions to 29, but the ensuing autumn temperature and rainfall frequencies were not different from the distributions to be expected from chance. On the other hand, progressive/southerly summers defined as (*a*) P_{45} S_{45} and (*b*) P_{345} S_{45} tended to be followed by warmer, wetter autumns. Anticyclonic/southerly summers also had a weak tendency to be followed by warmer autumns. Some relevant statistics are presented in Table IV.

TABLE IV—FREQUENCIES OF AUTUMN TEMPERATURE (QUINTILES) IN CENTRAL ENGLAND AND AUTUMN RAINFALL (TERCILES) OVER ENGLAND AND WALES FOLLOWING PROGRESSIVE, SOUTHERLY, ANTICYCLONIC SUMMERS AS SPECIFIED BY P, S AND C INDICES

Type of summer	Autumn Central England temperature					Autumn England and Wales rainfall			No.
	1	2	3	4	5	1	2	3	
(i) progressive/ southerly (*a*)	0	2	2	5	1	1	4	5	10
(*b*)	2	3	3	7	2	2	6	9	17
(ii) anticyclonic/ southerly (*a*)	2	4	5	5	4	7	9	4	20
(*b*)	3	5	7	8	6	10	10	9	29

Note : (i) (*a*) = $P_{45}S_{45}$; (*b*) = $P_{345}S_{45}$;
(ii) (*a*) = $C_{12}S_{45}$; (*b*) = $C_{123}S_{45}$.

None of the distributions shown in Table IV is statistically significant. However, item (i)(*b*) shows a trend in autumn temperature and rainfall which is almost complementary to the indications given in Tables I and II. A brief examination of autumn weather following those July plus August periods with progressive, southerly, anticyclonic characteristics did not suggest any significantly different frequency distributions from those involved in the summer selections of Table IV.

In the autumn months following the 17 occasions of $P_{345}\ S_{45}$ summers ((i)(*b*) of Table IV) no consistent sequence in temperature was in evidence; in each of the three months the frequency distribution is not statistically different from chance expectations, although it is noteworthy that only 3 of the 17 occasions were quintiles 1 or 2 in November. As regards monthly rainfall the frequency distributions are interesting. September showed no strong bias; in October tercile 2 occurred on 9 occasions (tercile 3 on 6); in November tercile 3 was observed on 10 occasions. Moreover, the four dry Septembers were all followed by wet Novembers. There is a suggestion that the tendency for above average rainfall following $P_{345}\ S_{45}$ summers is particularly a feature of late autumn whenever September is dry.

Conclusions. In this paper it has been shown that useful predictions of autumn weather, particularly seasonal mean temperature in central England, may be attempted whenever the preceding summer synoptic characteristics are suitable. The numerical parameters used for specifying the large-scale synoptic types over the British Isles are certain combinations of the *P*, *S* and *C* indices, which occur on rather more than 40 per cent of occasions. The main conclusions are briefly summarized below; in all cases temperature refers to central England, and rainfall to England and Wales.

(a) Blocked/northerly ($P_{12}\ S_{12}$) and/or blocked/cyclonic ($P_{12}\ C_{45}$) summers are generally followed by cold autumns.

(b) Cyclonic/northerly ($C_{45}\ S_{12}$) 'high summers' (i.e. July plus August) are generally followed by cold autumns.

(c) On some 25 per cent of occasions the synoptic classifications are $P_{12}\ S_{12}$ and/or $P_{12}\ C_{45}$ in summer and/or $C_{45}\ S_{12}$ in high summer, and on these occasions cold autumns are very likely to follow. As 40 per cent of autumns are cold (i.e. T_1 or T_2), predictions can be made on 25 out of 40 occasions.

(d) The synoptic pre-conditions given in *(a)*, *(b)* or *(c)* do not give a very reliable prediction of autumn rainfall, although average rainfall is most likely.

(e) Progressive, southerly, anticyclonic summers do not show similar reliability as predictors of warm autumns.

(f) Summers characterized by the synoptic parameters $P_{345}\ S_{45}$ (progressive/southerly) tend to be followed by warm rather than cold autumns and by wet rather than dry autumns. The data also indicate that above average rainfall is likely in late autumn (November), especially when September is dry.

Appendix

The quintile boundaries of the *P*, *S* and *C* indices for *(a)* summer (i.e. June + July + August) and *(b)* high summer (i.e. July + August) are based on the period 1874 to 1965. The quintile boundaries for each month of the year are given in the original paper.[3]

Quintiles		(a) Summer P	S	C	(b) High summer P	S	C
5/4	≥	63	8	20	58	7	21
4/3	≥	42	0	4	37	0	11
3/2	≥	24	−7	−9	18	−5	−3
2/1	≥	9	−17	−24	1	−14	−14

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RAPID CHANGES IN THE HEIGHT OF THE MELTING LAYER

By T. W. HARROLD, K. A. BROWNING and J. M. NICHOLLS

Summary. Two examples of rapid, advected, changes in the height of the melting layer in warm frontal precipitation are discussed. The examples include a case where two melting layers existed overhead simultaneously.

Introduction. In this country continuous rain is largely produced by the Bergeron process, in which precipitation forms at high altitudes and falls as snow to the 0°C level. The depth over which melting occurs below this level is a function of the temperature lapse and the nature of the particles. The melting layer is readily detectable by radar because the reflectivity of the wet snow is typically five times greater than that of the snow or rain immediately above or beneath the layer, thereby giving rise to the so-called radar 'bright band'. It is well known that the height of the melting level changes appreciably within frontal zones. The purpose of the present note is to describe two interesting examples in which these changes were concentrated into discontinuities.

Example 1 : Rapid descent of the melting layer in warm frontal precipitation. During the approach of a diffuse warm front on 21 August 1965, a vertically pointing radar at the Royal Radar Establishment, Pershore, showed a drop in the bright band from 3 to 2·2 km in less than 90 seconds.[1] Since the bright band subsequently remained near 2·2 km for at least an hour, it can confidently be related to the melting level rather than to the passage of a precipitation fallstreak as described by Browne.[2] The upper air sounding in Figure 1, thought to have been representative of the region east of the bright-band discontinuity, shows a dry subsidence layer between 2 and 3 km, just beneath the front. The initial bright-band level over Pershore corresponded closely to the 0°C level at the top of the subsidence layer in Figure 1. Because of the nearly isothermal lapse within the subsidence layer, only 1degC of cooling was needed to lower the 0°C isotherm to the level of the lower bright band at 2·2 km.

The sudden lowering of the bright band coincided with the arrival of the sharply-defined leading edge of an area of moderate to heavy precipitation at these levels. Evaporation and melting of the snow as it fell into the dry air beneath the original 0°C level is thought to have produced cooling in the

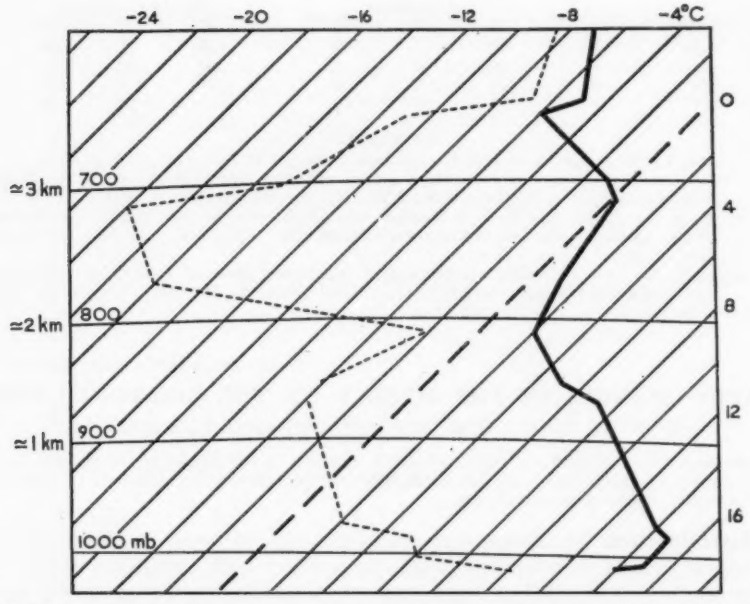

FIGURE 1—TEMPERATURE AND DEW-POINT PROFILES FROM HEMSBY AT 0000 GMT, 21 AUGUST 1965

Note the nearly isothermal layer between 2 and 3 km.

manner described by Findeisen[3] and Lumb.[4] The winds at these levels relative to the movement of the rain band were parallel to its leading edge, so that the cooled air was not advected away from the rain area, but rather remained within it long enough for the degree of cooling to become quite pronounced. As a result a mesoscale cold front developed, its passage being revealed by the sudden drop in the bright band.

Wexler[5] has shown that, of 10 occasions which he studied, the bright band dropped during the first hour or so of rain on 8 occasions; there was also evidence on other occasions that the bright band would lower sharply during moderate to heavy rain. The observation reported in this note is in keeping with his results. However the suddenness of the change in this example is greater than implied by Wexler and was due to a pre-existing deep temperature inversion near 0°C, in which the air was initially dry. The increased cooling within an area of moderate precipitation was sufficient to lower the 0°C level from the top to bottom of this inversion.

Example 2 : Rapid ascent of the melting layer in warm frontal precipitation. Although there is often a temporary lowering of the melting layer during the first hour or so of rain due to the cooling effects of evaporation and melting, the overall trend during the passage of a warm front is for the melting layer to ascend. Usually the ascent is gradual. However, during the passage of a well-marked warm front on 16 October 1967, there was a discontinuous jump from 1·1 to 2 km. (The airflow and precipitation structure

over England and Wales on this occasion have been described in detail by Browning and Harrold.[6]) Photographs (not reproduced) of the RHI display of an AN/TPS-10 3-cm radar at RRE, Pershore, showed a corresponding jump in the level of the bright band. Series of such photographs obtained at 10° azimuth intervals have enabled the horizontal extent of the bright-band discontinuity to be mapped at different times, as shown in Figure 2. The discontinuity was detectable for over an hour, the period being limited by the inability of the radar to detect slight precipitation at large ranges. Its movement is shown in Figure 2 to have been 10 m/s from 200°.

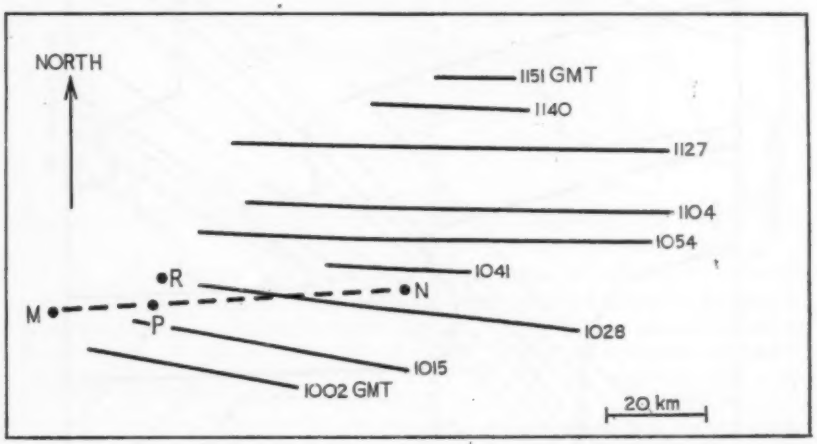

FIGURE 2—MAP SHOWING MOVEMENT OF THE MELTING-LEVEL DISCONTINUITY IN THE PERSHORE AREA

The shortening of the length of the discontinuity as it moved north is a result of the inability of the radar to detect slight precipitation at larger ranges. R denotes the position of a radiosonde passing through the upper melting level at 1045 GMT. P shows the position of Pershore, M of Malvern. MN shows the orientation of Plate I.

Plate I shows the RHI display of an AN/TPS-10 3-cm radar at RRE, Malvern, 20 km west of Pershore, and it illustrates the bright-band structure looking along the line MN of Figure 2 which was almost along the discontinuity at 1025 GMT as it passed over Pershore. This figure is a composite of two photographs, the one at closer ranges being obtained at a lower radar sensitivity so as to resolve features which otherwise would have saturated the display. Of particular interest is the fact that Plate I shows the two bright bands (at 1·1 and 2 km) to be overlapping over almost the entire range along MN. RHI photographs obtained at right angles to the bright-band discontinuity showed that the north–south extent of overlapping rarely exceeded 10 km.

Figure 3 shows a cross-section for 1200 GMT, 16 October 1967 through Pershore and perpendicular to the surface warm front. The section was constructed from routine radiosonde ascents together with four sequential ascents from Pershore displaced relative to the movement of the front. The heights of the two bright bands in Plate I correspond to the 0°C levels in the warm and cold air. As in the case of Example 1 in this paper, the discontinuity coincides with the passage of the near-isothermal layer at 0°C.

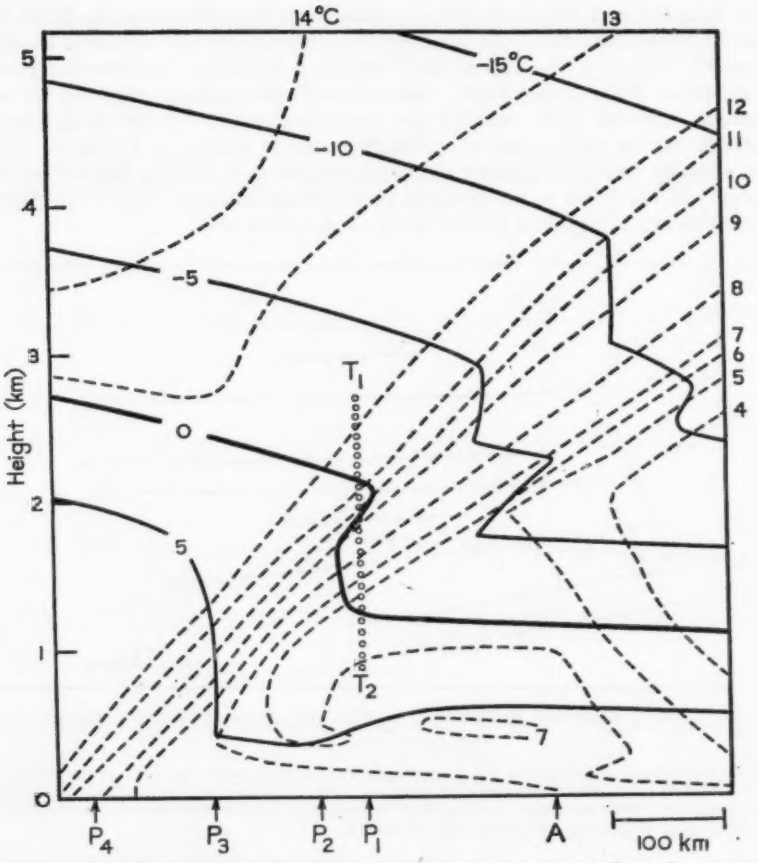

FIGURE 3—CROSS-SECTION THROUGH PERSHORE ALONG 200°–020°, PERPENDICULAR TO THE FRONT AT 1200 GMT, 16 OCTOBER 1967

———— Isotherms (°C) - - - - - - Wet-bulb potential temperatures (°C)

The arrows at the base of the figure show the positions of radiosonde ascents from Aughton (A) and Pershore (P), those at Pershore being displaced relative to the movement of the front.

Accurate measurements of wind and of precipitation fall-speed were made with a Doppler radar while the bright-band discontinuity passed over Pershore.[6] These measurements showed that, relative to the movement of the discontinuity, the wind was 250° 5 m/s at the upper bright band and 030° 5 m/s at the lower bright band, with almost zero relative velocity midway between. Mean precipitation fall-speeds in the vicinity of the discontinuity as measured by the Doppler radar are plotted in Figure 4*(a)*,* from which it can be seen that the mean fall-speed at the discontinuity was close to 3 m/s. Hence the horizontal displacement of particles would have been less than

* Divergence measurements by Doppler radar indicated vertical air motion of only 5 cm/s; therefore, the vertical velocities in Figure 4*(a)* are representative of the mean (reflectivity weighted) terminal velocities of the precipitation particles.

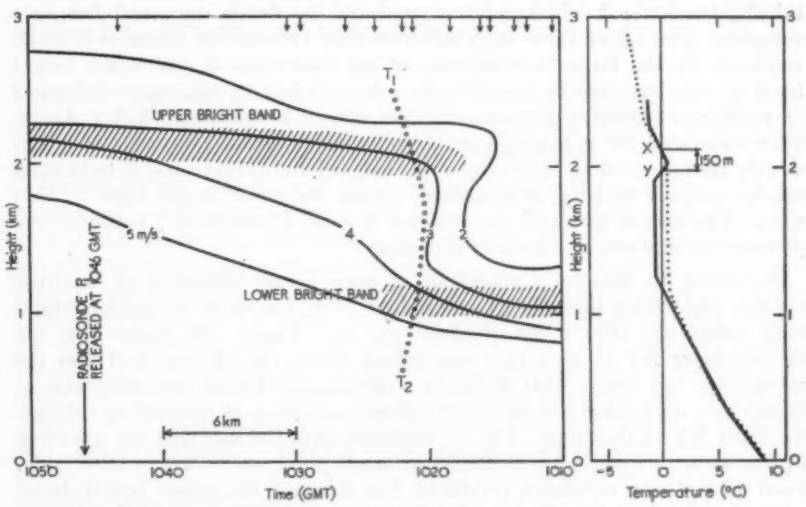

FIGURE 4*(a)*—HEIGHT-TIME SECTION OF MEAN FALL-SPEEDS OF PRECIPITATION OVER PERSHORE MEASURED WITH VERTICALLY POINTING DOPPLER RADAR

Arrows at top of diagram show times of vertical viewing. T_1ooooooT_2 is the trajectory of particles relative to the movement of the bright-band discontinuity.

FIGURE 4*(b)* — TEMPERATURE PROFILE NEAR PERSHORE FROM RADIOSONDES RELEASED AT 1046 GMT AND 1124 GMT

Note the shallow warm layer XY near 2 km, which was sufficient to produce complete melting of the snowflakes.

——— Radiosonde, P_1 in Figure 3 at 1046 GMT.

..... Radiosonde, P_2 in Figure 3 at 1124 GMT.

1 km in a fall between the two bright bands. This implies that precipitation falling through the upper bright band at about 1020 GMT will also have fallen through the lower bright band, along the trajectory T_1T_2 in Figures 3 and 4*(a)*.

The temperature sounding shown by the solid curve in Figure 4*(b)* is based upon a radiosonde released from Pershore at 1046 GMT (P_1 in Figure 3) which passed through the upper bright band at 1054 GMT in a position R in Figure 2, 10 km south of the bright-band discontinuity. For comparison, the ascent released at 1124 GMT (P_2 in Figure 3) is also shown. The occurrence of the two bright bands during the descent of particles along T_1T_2 can be explained on the basis of the solid curve in Figure 4*(b)*, by recalling that the radar reflectivity maximum within a bright band is due mainly to processes associated with the melting of snowflakes[7]. The explanation is in three stages. Firstly, the upper bright band was produced by the *partial* melting of snowflakes as they descended through the layer corresponding to XY in Figure 4*(b)*. Secondly, the resulting wet snowflakes, instead of melting into raindrops, refroze as they descended into colder air below the upper bright band. And

thirdly, the lower bright band was produced by the further, and this time complete, melting of these snowflakes as they crossed the lower 0°C level. Evidence for the incomplete melting of the snowflakes at the upper bright band at 1020 GMT can be found in the relatively low (3 m/s) mean fall-speed of particles descending between the two bright bands along T_1T_2. Later, after 1050 GMT for example, particles attained a mean fall-speed of 5 m/s shortly after descending below the upper bright band (Figure 4(*a*)), indicating nearly complete melting of snowflakes within the upper bright band at that time. The ascent at 1124 GMT (dotted line in Figure 4(*b*)) confirms the presence of only one 0°C level at that time.

According to Mason[8] a snowflake of mass 1 mg comprised of dendritic crystals and falling through an atmosphere with a lapse of 6degC/km would melt completely after a fall of about 150 m. Figure 4(*b*) shows that the melting layer XY at 1054 GMT was indeed about 150 m deep and that the mean lapse was close to that in Mason's calculation. Hence it is reasonable to expect that snowflakes will have just melted completely in descending through the layer XY at that time. This is consistent with the fact that the sounding P_1 was made in a region where the lower bright band was no longer detectable. Even though the raindrops produced just beneath the upper bright band at 1054 GMT might have become frozen during their descent below 2 km, the subsequent melting of these ice pellets would not be expected to have produced a second bright band (owing to the absence of major fall-speed and drop break-up effects).

Conclusions. The sudden changes in the height of the melting layer reported here were due to the existence of isothermal layers near the 0°C isotherm within or just beneath the frontal zone. Isothermal layers are not unusual in fronts and, therefore, such sudden changes may not be an uncommon feature.

Complete melting of snow has been shown in one example to occur within a layer only about 150 m deep in which the mean temperature was about 0·4°C. When an even shallower layer of this kind was embedded within air colder than 0°C, it was found that the snow melted only partially, and then froze again on descending further. In this circumstance a double radar bright band was produced.

Acknowledgement. This paper is published with the permission of the Director of the Royal Radar Establishment.

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551.513.1:551.547.5

MIDDLE-LATITUDE WAVELENGTH VARIATIONS AT 500 mb

By P. D. DE LA MOTHE

Summary. The wavelengths represented by some of the major 500 mb troughs and ridges are examined, and how these wavelengths varied throughout the year in latitude 50°N during the period 1949–64 is considered. The trough-ridge or Hovmöller diagram is described, and graphs are drawn showing variations of average wavelength with time.

Introduction. The purpose of the current work is to present data relating to seasonal wavelength variations of some of the main climatological features at 500 mb around the hemisphere at latitude 50°N. The basic data used, which were derived from German punched cards and transferred to magnetic tape, consisted of 500 mb contour heights for the period 1949–64. These were arranged so that a direct computer print-out of mean contour heights for each pentad (5-day period) during each year was obtained making possible the construction of a trough-ridge diagram, similar to that originally devised by Hovmöller.[1] Briefly, this is a diagram on which 500 mb contour heights, for a given latitude around the hemisphere, are plotted as functions of longitude and time. Various isopleths and different types of shading or hatching may then be employed to show the intensity of troughs and ridges, and their seasonal progression.

Method. In the present work, trough-ridge diagrams for each year of the period 1949–64 were analysed by simply selecting maximum and minimum contour-height values for each pentad around the hemisphere, and attempting to delineate the continuity or otherwise of the main features throughout the year from one pentad to the next. Figure 1 shows part of the trough-ridge

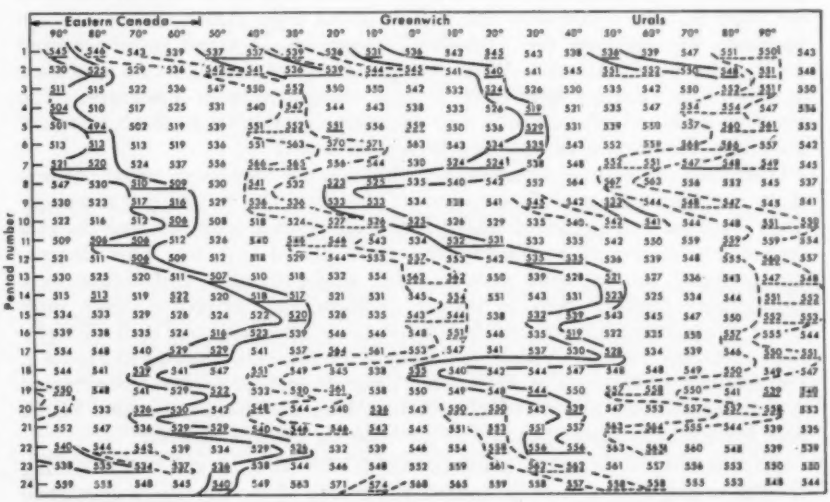

FIGURE 1—SECTION OF TROUGH-RIDGE OR HOVMÖLLER DIAGRAM SHOWING 500 mb PENTAD MEAN CONTOUR HEIGHTS DURING PERIOD JANUARY TO APRIL 1963, AROUND 50° NORTH

Contour heights are in geopotential decametres.

———— minimum value (trough axis) – – – – – maximum value (ridge axis)

diagram for 1963, and covers the first four months of that year, i.e. up to and including pentad 24, for the sector 90°W to 100°E. The original diagrams cover the sectors from 150°W to 170°E. (No data were available for the central Pacific Ocean during the period studied, so a full hemispheric coverage was not possible.) In Figure 1 the troughs and ridges emerge quite clearly, troughs being indicated by continuous lines and ridges by pecked lines. It is apparent from Figure 1 that apart from the major features, there are often intermediate or minor troughs and ridges. One example can be seen in pentads 9 and 10 between longitudes 30°E and 60°E. Furthermore these minor features increase in number and complexity during the summer. This is a natural outcome of the fact that there are more troughs and ridges on pentad charts throughout the year, but especially in summer, than on the longer-term climatological charts. However, particular attention was given to the sector from the Rocky Mountains past the Greenwich meridian to eastern Siberia, and the main 500 mb features studied were (i) the eastern North American or so-called 'Canadian' trough, (ii) the Atlantic ridge, (iii) the European trough, and (iv) the western Siberian ridge.

The measurement of wavelength is expressed throughout in degrees of longitude; the measurement for troughs is made from ridge to ridge either side of the trough, and for ridges is made from trough to trough either side of the ridge. In this context, the wavelength refers to the spacing at 50°N across the axis of the wave, expressed in degrees of longitude.

This procedure was carried out subjectively for each pentad of each year from 1949 to 1964, trough-ridge diagrams for each year having been previously derived and analysed as described above. One difficulty which arises in this method is the choice of criteria to be applied in order to identify correctly the main features when they are well removed from their usual or average positions. For example, the 'Canadian' trough can be found anywhere from the Rockies to mid-Atlantic. When it moves eastwards and takes up a position in, say, mid-Atlantic, invariably a new trough appears upstream. The question arises, which trough is now correctly identifiable as the 'Canadian' trough for the purpose of wavelength measurement ?

Obviously the first requirement is to find out the average or usual position of the main features at different times of the year, and to this end a trough-ridge diagram was prepared which indicated pentad mean contour heights expressed as an overall average for the whole period 1949–64. This diagram was analysed in the usual way and the major troughs and ridges were considerably clarified both as regards position and continuity, since most of the minor perturbations and oscillations were smoothed out in the longer-term averaging process. A simplified form of the diagram is shown in Figure 2. The required wavelength on each year's diagram was then taken to be related to that feature, the axis of which was closest to the average position of the same feature in the same pentad, on the 1949–64 average diagram.

Occasionally, a situation arose in which the axes of two neighbouring troughs or ridges were equally spaced either side of the average longitudinal position. If these two waves were completely symmetrical, then obviously either wavelength was acceptable, the matter of identification being purely academic.

If, on the other hand, the waves were not completely symmetrical, then it was necessary to consider the element of continuity. In fact, throughout

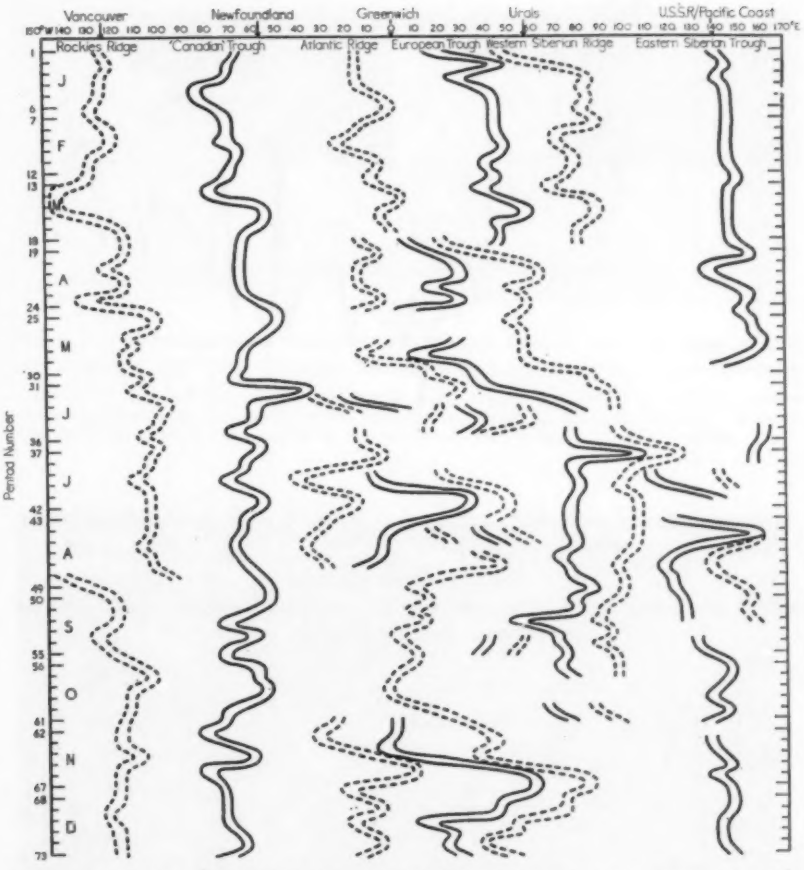

FIGURE 2—500 mb PENTAD MEAN TROUGH-RIDGE DIAGRAM, 1949 TO 1964 AVERAGE AROUND POSITIONS 50° NORTH

the entire process of identification of the main features, and particularly in cases of doubt, continuity was considered to be of paramount importance and was kept to the forefront in the application of subjective judgement.

Wavelengths of the main fatures having been derived in this way, for each pentad of the individual years under consideration, the average pentad wavelength for the period was calculated, and graphs drawn showing wavelength as a function of time for each of the four features (Figures 3, 4, 5 and 6).

Discussion. The most immediately obvious feature of the graphs is the shortening of wavelength from winter to summer. In winter, the wavelengths across the 'Canadian' trough, the Atlantic ridge and the western Siberian ridge (but *not* the European trough), average between 90 and 110 degrees of longitude, implying something near a 4- or 3-wave system around the hemisphere. In the summer, however, wavelengths average between 55 and

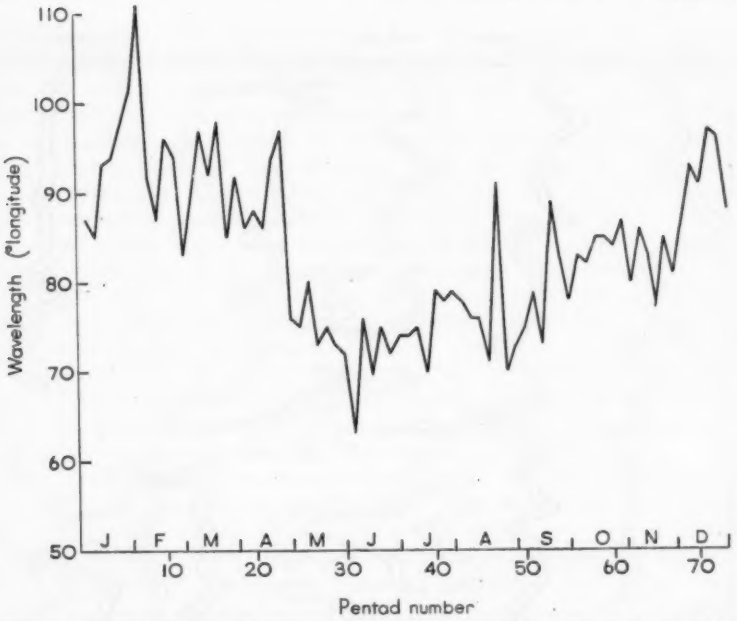

FIGURE 3—AVERAGE WAVELENGTH ACROSS 'CANADIAN' TROUGH (500 mb) AT 50° NORTH. PERIOD 1949 TO 1964

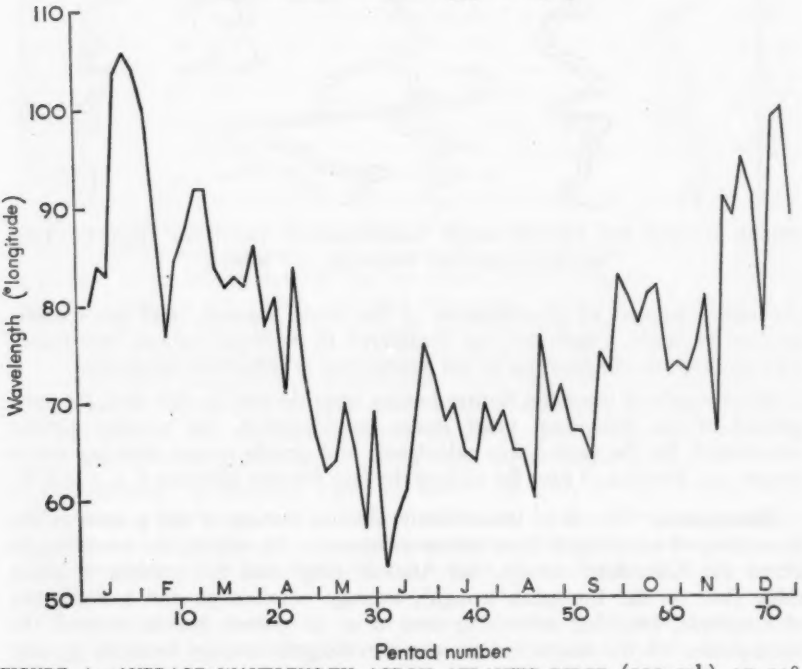

FIGURE 4—AVERAGE WAVELENGTH ACROSS ATLANTIC RIDGE (500 mb) AT 50° NORTH. PERIOD 1949 TO 1964

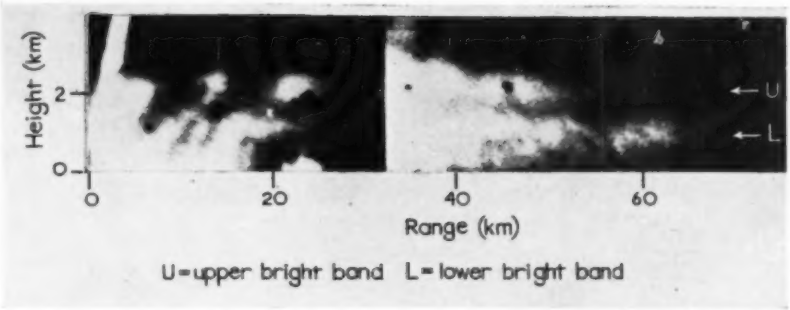

PLATE I—PHOTOGRAPH OF MALVERN RHI RADAR DISPLAY AT 1029 GMT ALONG THE AZIMUTH 087° (MN) SHOWING BRIGHT BANDS CO-EXISTING AT TWO HEIGHTS

The radar does not have any correction for range attenuation and so in order to resolve features at close range the portion of the RHI display within 32 km range has been obtained with the radar receiver less sensitive by a factor of 10. (See p. 329).

Photograph by courtesy of the Australian News and Information Bureau

PLATE II—PART OF THE COMPUTER SYSTEM (IBM 360/65) INSTALLED IN JULY 1968 AT THE BUREAU OF METEOROLOGY, MELBOURNE

The computer will enable the Bureau to play its part in the World Weather Watch as a World Meteorological Centre collecting and disseminating weather information from the southern hemisphere. It will soon be augmented by another computer which will be used to ensure that an unbroken 24-hour service is maintained and to carry out research on new weather forecasting techniques.

Photograph by courtesy of Mr Arthur Brühlmeier

PLATE III—SNOW CYLINDERS AT GELTWIL, SWITZERLAND

See page 350

Photograph by courtesy of Mr Arthur Brühlmeier

PLATE IV—SNOW CYLINDERS AT GELTWIL, SWITZERLAND

See page 350

To face p. 337

Photograph by courtesy of the Building Research Institute

PLATE V—FROST DAMAGE DURING BUILDING OPERATIONS

The mortar has not set properly during a spell of frost in spring 1959 in the Watford area and the brick wall has been pushed over.

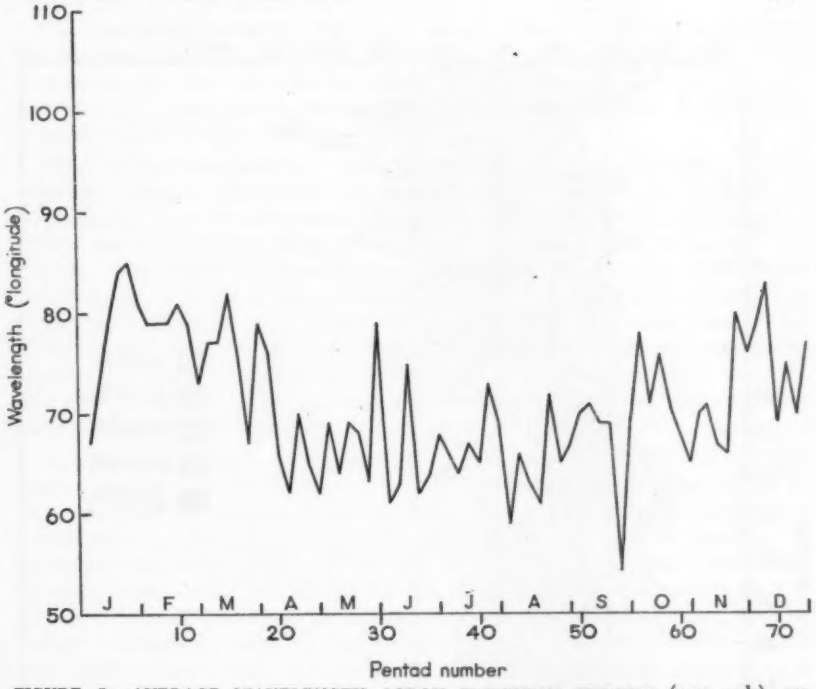

FIGURE 5—AVERAGE WAVELENGTH ACROSS EUROPEAN TROUGH (500 mb) AT 50° NORTH. PERIOD 1949 TO 1964

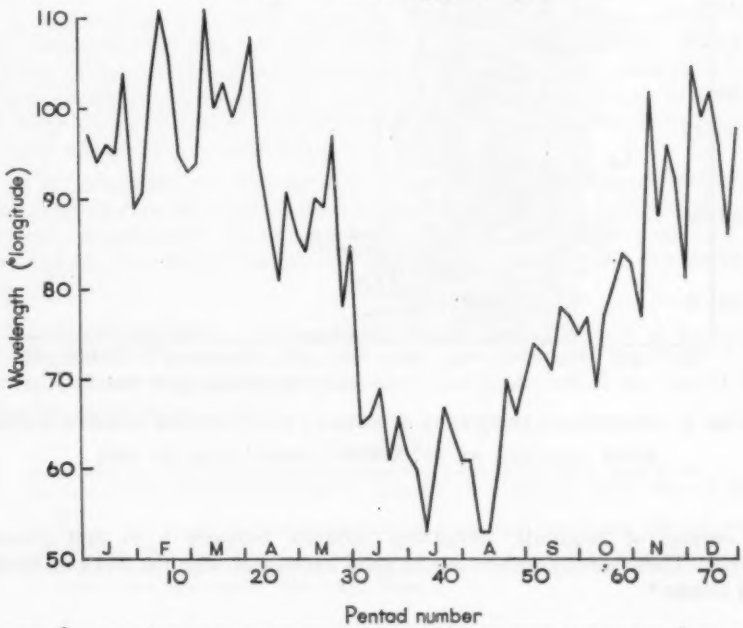

FIGURE 6—AVERAGE WAVELENGTH ACROSS WESTERN SIBERIAN RIDGE (500 mb) AT 50° NORTH. PERIOD 1949 TO 1964

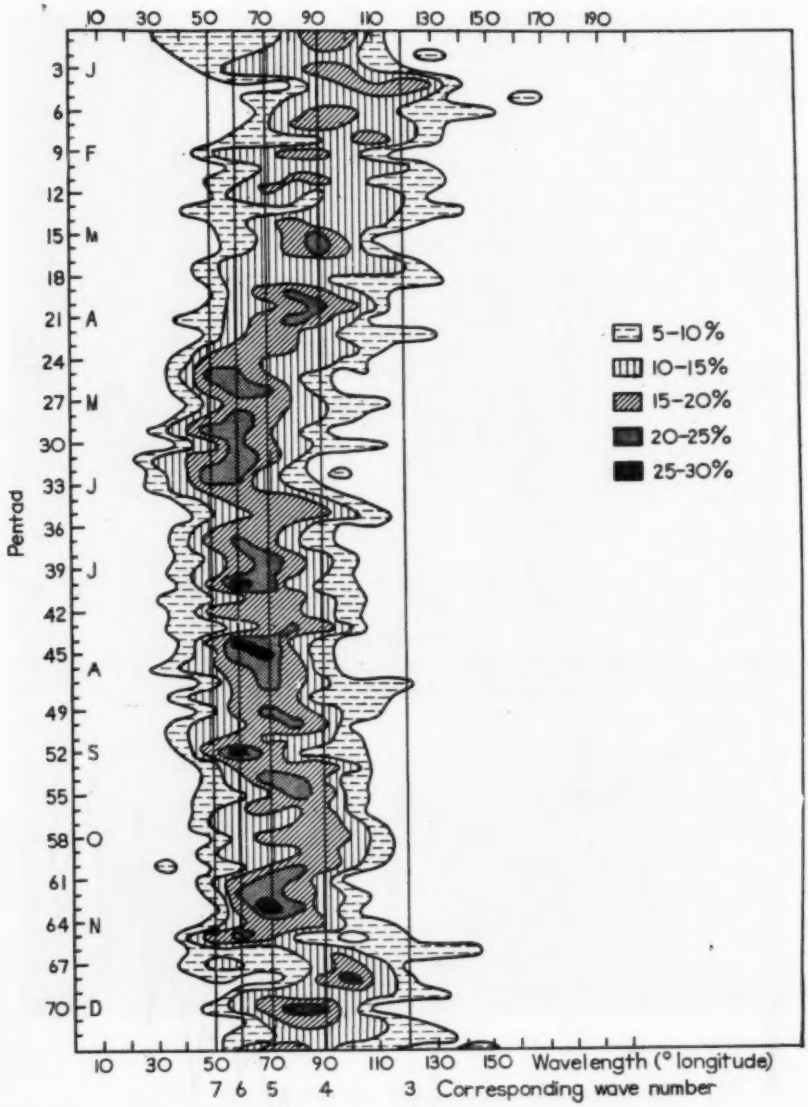

FIGURE 7—PERCENTAGE FREQUENCY OF VARIOUS WAVELENGTHS ACROSS ATLANTIC RIDGE (500 mb) AT 50° NORTH. PERIOD 1949 TO 1964

75 degrees of longitude, indicating variation between a 7- and 5-wave system. These wave numbers are in good agreement with the work of Riehl[2] and Glaser.[3]

The wavelength across the European trough (Figure 5) shows considerably less seasonal variation than the three other neighbouring features. It is noticeable also that the absolute range of wavelength is a great deal less than that of its neighbours, varying from around 80–85 degrees in midwinter to 55–60 degrees in late summer.

These points would seem to underline the fact that the European trough, although a major influence in the British Isles region, must be regarded as a subsidiary wave in the overall hemispherical wave system — the two most intense waves, of course, being the 'Canadian' trough and the eastern Siberian trough. Nevertheless, it must be emphasized that as far as the British Isles is concerned, a study of the behaviour of the European trough is becoming increasingly fruitful with regard to longer-term or seasonal forecasting. A correlation between the April European trough position and the ensuing summer rainfall in England and Wales has been demonstrated by Ratcliffe and Collison.[4] However from wavelength variations alone, no clear relationships are immediately apparent.

Another characteristic of the 500 mb troughs and ridges as shown by these pentad wavelength variations, is the tendency for certain wavelengths to occur in spells or régimes. There is, generally speaking, a spell of maximum wavelengths in December, especially early December, and in January. A systematic shortening takes place from about February onwards until May to early June. How this affects the whole sample of years may be seen in Figure 7 showing frequency of wavelengths across the Atlantic ridge. Slightly longer wavelengths occur in late June and affect July and August. There is a systematic increase of wavelength through the autumn apart from a peculiar, and evidently important, discontinuity associated with blocking in November.

It is noticeable once again when the European trough is examined, that the delineation of these régimes is not quite so clear. Furthermore, the western Siberian ridge shows a slight complication during the period of shortening wavelength in April and May since there appears to be a feint at a return to longer wavelengths in mid-May before finally settling to shorter wavelengths in late May to early June.

It is reasonable to suppose that these seasonal wavelength variations, and possibly even the changes from pentad to pentad, may be connected with climatic singularities and/or circulation changes in various sectors of the hemisphere, but the precise relationships are not entirely clear at the present time.

Acknowledgement. This presentation of data represents a temporary diversion from a major investigation under the direction of Mr H. H. Lamb of the seasonal progression of the 500 mb circulation, and I am indebted to him for guidance and advice.

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551.509.314:551.511.32

SOME NOTES ON BAROCLINIC INSTABILITY

By A. E. PARKER

Summary. The results of Fleagle's[1] theoretical work on baroclinic instability[1] is presented in a form which can be readily applied by the practical forecaster. The role of horizontal cross-wind shear in assisting vertical wind shear to produce baroclinic instability is also explained. Finally a graph is presented which enables the deepening of a depression in 24 hours to be estimated using well-known parameters.

Introduction. In order to obtain greater insight into the day-to-day development of 500 mb patterns, with short- and medium-range forecasting in mind, it was decided to examine Fleagle's[1] theoretical work on this in mind, it was decided to examine Fleagle's (1957) theoretical work on this subject. The object of the present paper is to present Fleagle's results, and the present writer's results derived from them, in a form suitable for use by a practical forecaster. By presenting these results in a form which can be readily understood, the basic fundamentals of development should become clear to the student of meteorology, and the results of the test of the theory with regard to the deepening of depressions should be of use to the practical forecaster.

Baroclinic instability and the critical vertical wind shear. A long zonal wind current is said to be baroclinically unstable if a small quasi-horizontal disturbance in the wind current amplifies with time. Fleagle, using perturbation theory for the motion of air on the stream surface of maximum slope, obtained a criterion for the onset of baroclinic instability. His results are strictly applicable only to this stream surface, but in practice should apply to a layer of considerable depth in the middle troposphere. The complete derivation of Fleagle's criterion will not be given here, but the following brief account should assist in understanding the basic dynamics involved.

Fleagle wrote down the equations of motion for a broad zonal current of air, having a velocity U and located in the middle troposphere ,when given small perturbations of velocity u, v and w in the zonal, meridional and vertical directions respectively; u, v and w were then assumed to be given by,

$$\begin{aligned} u &= A \exp i\,(m\lambda + n\theta - \sigma t) \\ v &= B \exp i\,(m\lambda + n\theta - \sigma t) \\ w &= C \exp i\,(m\lambda + n\theta - \sigma t) \end{aligned} \qquad \ldots (1)$$

in which A, B and C are the amplitudes of the u, v and w perturbations respectively, λ the longitude, θ the co-latitude, t the time and m, n and σ constants. The values of u, v and w (w is of course zero when the motion is referred to the stream surface) given by equations (1) were then substituted into the equations of motion. The conditions of the problem enable σ to be found and baroclinic instability exists if σ becomes imaginary since this means that the perturbation increases with time. Fleagle's criterion for baroclinic instability is

$$l\,\frac{\mathrm{d}U}{\mathrm{d}z} > 2\left(\frac{(\omega + K)\sin\theta}{a}\right)\left(\frac{s_z}{l\gamma_i} + \frac{g\gamma l}{\alpha^2}\right)$$

in which the symbols have the following meanings;

θ the co-latitude

ω the scalar angular velocity of the earth

K the angular velocity of the air about the earth's axis relative to the earth's surface

l $2(\omega + K)\cos\theta$

U the unperturbed zonal wind and assumed by Fleagle to be given by $aK\sin\theta$

z the height

a the earth's radius

ρ the density of the air

γ $1/c^2$

c the speed of sound in air

s_z $-\left(\frac{1}{\rho}\frac{\partial\rho}{\partial z} + g\gamma\right)$

g the acceleration due to gravity

α^2 $\dfrac{m^2 - \cos^2\theta + n^2\sin^2\theta}{a^2\sin^2\theta}$

It can be shown that

$$-\left(\frac{1}{\rho}\frac{\partial\rho}{\partial z} + g\gamma\right) = \frac{1}{T}\left(\frac{dT}{dz} + \Gamma\right)$$

where T is the temperature at the level under consideration, dT/dz is the lapse rate at this level and Γ is the dry adiabatic lapse rate. For convenience in practical use the mean value of $(1/T)(dT/dz + \Gamma)$ for the layer sea level (or surface) to 500 mb will be used since the parameters required are readily available from charts. On making these substitutions and using φ the latitude instead of θ the co-latitude, Fleagle's criterion for baroclinic instability becomes

$$\frac{dU}{dz} > \frac{c^2}{2}\frac{1}{T}\left(\frac{dT}{dz} + \Gamma\right)\frac{\cos^2\varphi}{\sin^2\varphi\,(\omega a\cos\varphi + U)} + \frac{2g}{c^2}\left(\frac{\omega a\cos^2\varphi + U\cos^2\varnothing}{m^2 - \sin^2\varphi + n^2\cos^2\varphi}\right) \quad \ldots (2)$$

The criterion (expression 2) gives the minimum value of vertical wind shear dU/dz, assumed constant with height, which must exist before baroclinic instability can occur and may be termed the critical shear. Fleagle also obtained a complicated expression for the rate of amplification of a disturbance with time and he summarized his results in Figure 1. It can be shown that a temperature gradient of 2 sinφcosφdegC/(deg/lat.) (Figure 1) corresponds to a vertical wind shear of about 25 kt in 4880 m (i.e. say from 900 m to 500 mb approx.). Figure 1 shows that baroclinic disturbances amplify most rapidly at latitude 60° with a zonal wavelength of about 4000 km (or zonal wave numbers m of from 4 to 6) and assuming a constant meridional wave number n of 4. Another important result emerging from the theory is that when a trough-ridge system amplifies so does the vertical motion which is usually upwards ahead of a trough and broadly downwards in the ridge.

Values of dU/dz, expressed as wind shear in knots per 4880 m (16·000 ft, i.e. the shear from 2000 to 18 000 ft, a parameter frequently used by forecasters), were calculated from the right-hand side of expression (2) using the Meteorological Office computer and assuming a series of values for m, n and dT/dz and the results are given in Figure 2. It may be helpful when using Figure 2 to remember that a value for $(1/T)(dT/dz + \Gamma)$ of $1{\cdot}0 \times 10^{-3}\text{ft}^{-1}$

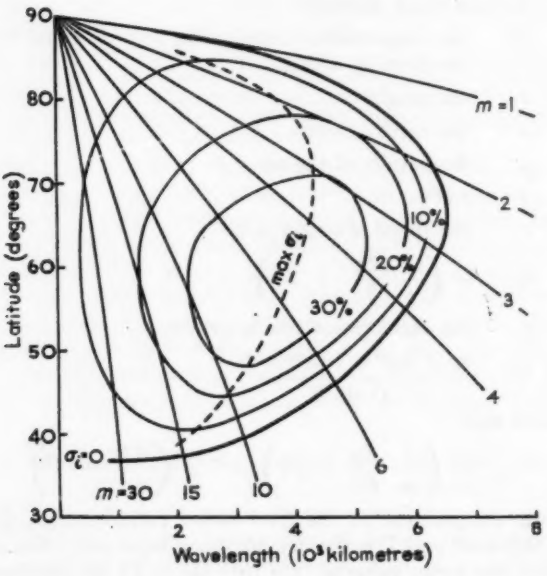

FIGURE 1—DISTORTION OF 500 mb FLOW PATTERN

Amplification in 24 hours (labelled in percentage) as a function of wavelength (or zonal number) and latitude, for the following values of the parameters: latitudinal temperature gradient, 2 sin φ cosφ degC/degrees of latitude; temperature, 260°K; lapse rate, 6·5deg K/km; $K = 0{\cdot}6 \times 10^{-6}$/s. σ_i is percentage amplification in 24 hours. ($n = 4$).

This figure is reproduced by courtesy of the Royal Meteorological Society.

($0{\cdot}308 \times 10^{-5}\text{m}^{-1}$) corresponds closely to isothermal conditions while a value of zero corresponds to the dry adiabatic lapse rate, or to cloudy air having a lapse rate within it equal to the saturated adiabatic lapse rate in which case Γ in $(1/T)(\mathrm{d}T/\mathrm{d}z + \Gamma)$ must be replaced by β the saturated adiabatic lapse rate. It is important to note that the right-hand side of expression (2) does not become zero for zero static stability.

Each line in Figure 2 corresponds to a particular latitude and zonal wave number m both of which are shown on the lines. For example, considering the line for latitude 50° and zonal number 4, in order to produce baroclinic instability with a static stability of $0{\cdot}5 \times 10^{-5}\text{ft}^{-1}$ ($0{\cdot}154 \times 10^{-5}\text{m}^{-1}$) for the layer sea level to 500 mb, the vertical wind shear must exceed 29 kt in 4880 m (16 000 ft).

The critical latitude. It is clear from Figure 2 that, for constant static stability, the vertical wind shear necessary to produce baroclinic instability *increases* with *decreasing* latitude. Thus when baroclinic instability exists in middle or high latitudes there will always be a latitude south of which baroclinic stability exists. This latitude is called the critical latitude and clearly its position can vary considerably from day to day.

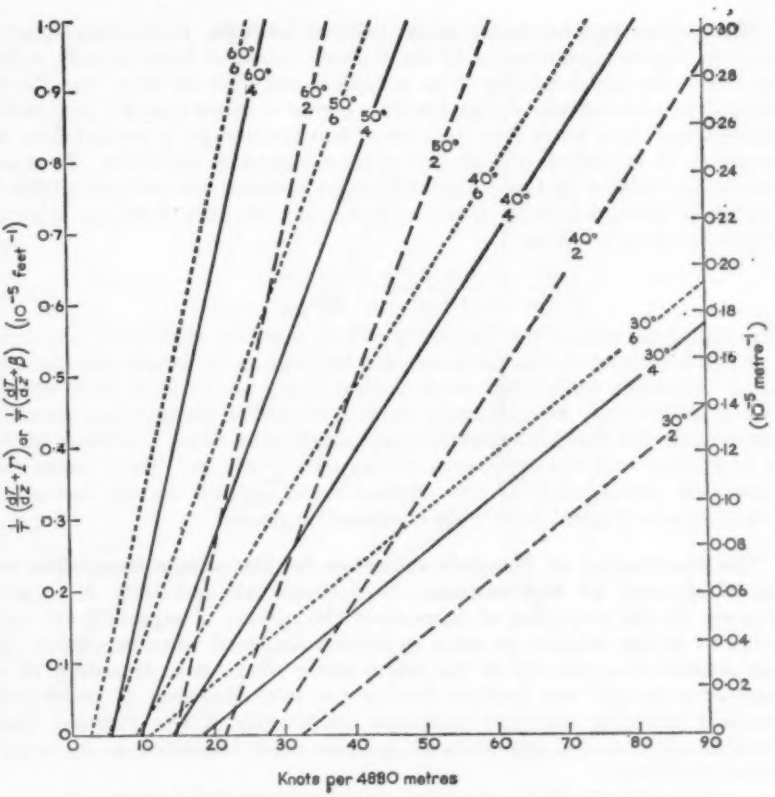

FIGURE 2—VERTICAL SHEAR (KNOTS PER 4880 METRES)

Cut-off vortices. Figure 1 shows that the amplitude of baroclinic disturbances must vanish at the poles ($\sigma_i = 0$) and at the critical latitude. At latitude 80° only about two amplifying troughs are theoretically possible round the hemisphere while at latitude 50°, for example, disturbances of zonal wave number 8 amplify most rapidly. It follows that only about two amplifying troughs can extend from latitude 80° to 50° and any other amplifying troughs at latitude 50° tend to become cut-off vortices.

Formation of jet-streams. Considering latitude 30°, for a zonal wave number of say 4 and a static stability of $0{\cdot}5 \times 10^{-5}$ft^{-1} ($0{\cdot}154 \times 10^{-5}$ m^{-1}) then from Figure 2 a vertical wind shear of 81 kt in 4880 m is required to produce baroclinic instability. The corresponding value at latitude 50° is only about 29 kt so that the neighbourhood of latitude 30° tends to be a baroclinically stable region. It follows that, since troughs to the north of the critical latitude can amplify while those to the south cannot, jet streams will at least be possible to the south of the 500 mb trough and near the critical latitude. Jet streams can of course form in other places besides near the critical latitude, for example at the crest of the ridge because of the crowding together of isotherms produced by a confluent wind field.

The region equatorwards of the critical latitude. It has been deduced that the region equatorwards of the critical latitude is baroclinically stable for lapse rates which usually occur in middle and high latitudes, but it is of interest to examine this region further. Figure 2 shows that for zero static stability and for a zonal wave number of 6 at latitude 30° a vertical shear of about 11 kt in 4880 m should just produce baroclinic instability. Now for zero static stability in low latitudes Fleagle's criterion for instability (which cannot be derived directly from expression (2) but only from the original expression for σ) becomes

$$\frac{dU}{dz} > \frac{2ag\ (\omega + K)\ \cos^3\varphi}{c^2\ (m^2 + n^2\cos^2\varphi)} \qquad \ldots (3)$$

The right-hand side of this inequality will be small for negative K, i.e. anticyclonic horizontal cross-wind shear, and for large m, i.e. small-scale disturbances. Therefore easterly waves or tropical storms are liable to form where, over a limited area, the hydrostatic stability is small or negative (i.e. unstable cloudy air), and there is horizontal anticyclonic wind shear, conditions likely to be fulfilled over the ocean between latitudes 5° and 20°, but of course the subsequent development of any tropical storm depends on the release of large amounts of latent heat in the condensation process.

The application of Fleagle's criterion for baroclinic instability to the deepening of depressions. Since baroclinic instability is a prerequisite for the deepening of depressions, the validity of expression (2) was tested by noting whether or not a depression deepened when expression (2) was satisfied over the tip of the warm sector of a wave depression in a reasonably straight and uniform flow at 500 mb. However, in order that practical examples were not restricted too severely it was assumed that Fleagle's criterion also applied to air currents when orientated by up to 45° either side of zonal.

Sixty-six depressions, some of which did not deepen, which occurred during October, November and December 1957, and during January, February and December 1958 were examined, the data being extracted from *Daily Series Synoptic Weather Maps, Part* I.[2] The static stability was computed in the warm air at the tip of the warm sector using the dry-bulb temperature at 500 mb, the surface temperature, and the height of the 500 mb surface from the 500 mb contour chart. The vertical wind shear was obtained by reading off the 500 mb wind over the tip of the warm sector and then using the formula

$$\text{shear in 4880 m} = \frac{U \times 4880}{H}$$

where U is the 500 mb wind over the centre and H the corresponding 500 mb contour height. If the surface wind over the tip of the warm sector was not calm, then an allowance was made for this by using the wind shear vector from 900 m to 500 mb and not the actual wind at 500 mb.

A practical example to illustrate the method. To illustrate the method consider Figures 3*(a)* and *(b)*, the sea-level and 500 mb charts for 1200 GMT for 21 February 1958. The surface chart Figure 3*(a)* shows a wave on a cold front at about 33°N 47°W, the wave occurring under a strong

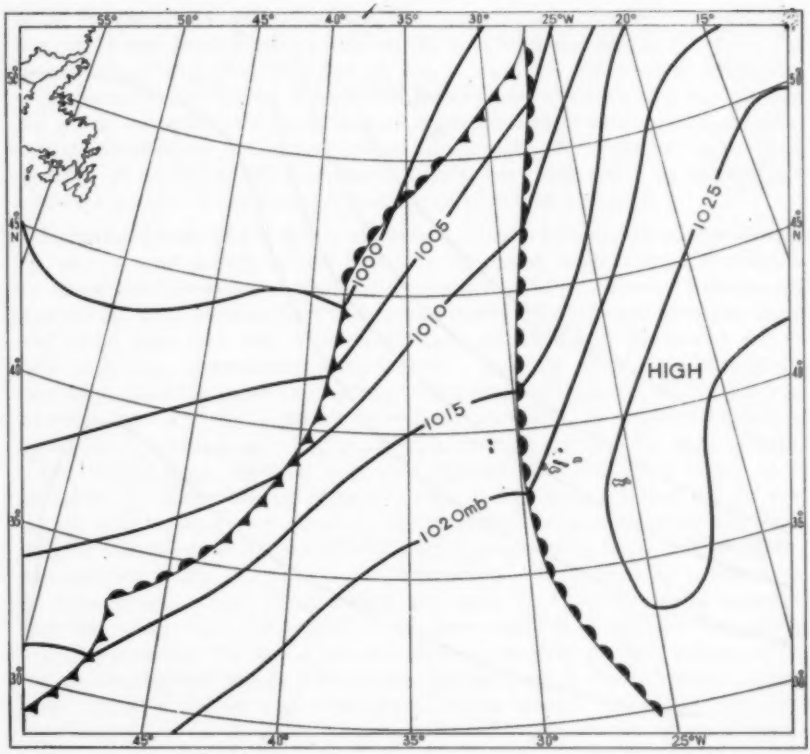

FIGURE 3*(a)*—SEA-LEVEL CHART, 1200 GMT, 21 FEBRUARY 1958

flow at 500 mb as shown in Figure 3*(b)*. The temperature at the surface was taken as 21°C, and that at 500 mb as – 18°C which, combined with a contour height over the wave of 5680 m, gave a mean lapse rate from the surface to 500 mb of 2·09 degC per 1000 ft (or ·0686degC per 100 m). This is rather a large lapse rate but it may be noted in passing that deepening depressions in low latitudes usually have unstable air in their warm sectors. The lapse rate of 2·09 deg C per 1000 ft yielded the rather low value of $0{\cdot}33 \times 10^{-5}\,\mathrm{ft}^{-1}$ ($0{\cdot}1 \times 10^{-5}\mathrm{m}^{-1}$) for the stability factor $(1/T)(\mathrm{d}T/\mathrm{d}z + \Gamma)$. The vertical wind shear from the surface to 500 mb was taken as 50 kt, which gives a shear of 43 kt in 4880 m. Now from Figure 2, at latitude 33° and for a static stability of $0{\cdot}33 \times 10^{-5}\mathrm{ft}^{-1}$, a vertical wind shear of 47 kt in 4880 m is required to produce baroclinic instability in non-cloudy air, a value somewhat greater than the value estimated as being present. During the next 24 hours this wave might have been expected to travel north-eastwards at a mean speed of say 30 kt (the wind speed at 700 mb), and thus be located at or near latitude 45°. The mean latitudinal position during the 24 hours would be 1/2 (33° + 45°) = 39°, and at this latitude a shear of 33 kt per 4880 m would be required to produce baroclinic instability in cloudless air. Hence the wave at 33°N on 21 February 1958 would be expected to move north-eastwards and soon

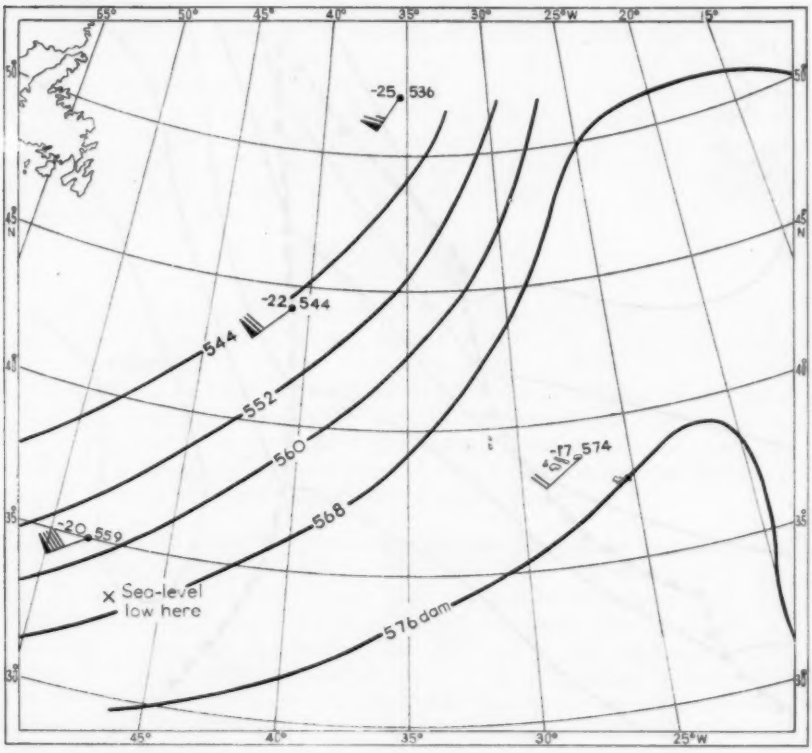

FIGURE 3(*b*)—500 mb CHART 1200 GMT, 21 FEBRUARY 1958

commence to deepen quite rapidly. The amount of deepening can be estimated by anticipating later results given in Figure 4. Since thick alto-stratus was reported on the cold front the static stablity was taken as zero (it was probably negative), and from Figure 2 at latitude 39° a shear of only

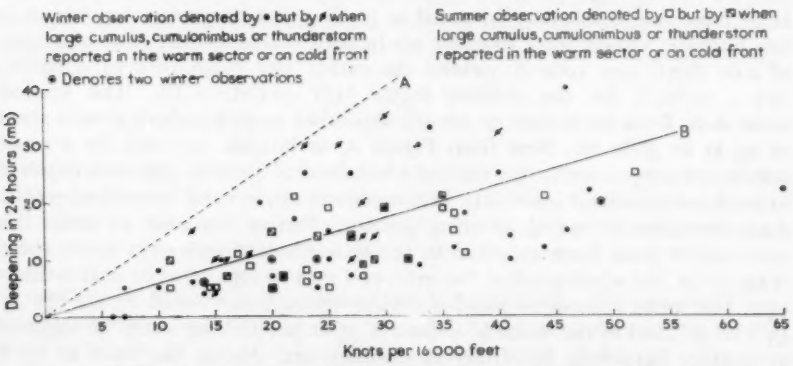
Winter observation denoted by • but by ⧸ when large cumulus, cumulonimbus or thunderstorm reported in the warm sector or on cold front
Summer observation denoted by □ but by ⧄ when large cumulus, cumulonimbus or thunderstorm reported in the warm sector or on cold front
• Denotes two winter observations
A
B
Deepening in 24 hours (mb)
Knots per 16 000 feet

FIGURE 4—EXCESS SHEAR (KNOTS PER 16 000 FEET 488om)

10 kt in 4880 m would give baroclinic instability. Thus the excess shear was 33 kt and from Figure 4 this would lead to a deepening of 19 mb for average deepening (line XB) but 43 mb in 24 hours for extreme deepening in unstable warm sectors (line XA). Some large cumulus was reported in the warm sector but the cloud type was mainly stratocumulus, and cumulus and stratocumulus, so it seems reasonable to take the mean of 19 mb and 43 mb, i.e. 31 mb as the deepening of this wave depression in 24 hours, a result quite close to the actual deepening of 28 mb in 24 hours.

Statistical test of Fleagle's criterion. Of the 66 depressions considered, 17, which were mostly in low latitudes, deepened when Fleagle's criterion for baroclinic instability in *cloudless* air was not satisfied. These 17 deepening depressions were re-examined and it was found that on each occasion there was cloud and rain and often hydrostatic instability in the warm sector. Now it follows immediately from Figure 2 that, for cloudy air having the saturated adiabatic lapse rate within it, the static stability $(1/T)(\mathrm{d}T/\mathrm{d}z + \beta)$ becomes zero and the vertical wind shear necessary to produce baroclinic instability becomes small. For example, at latitude 50° and for wave number 4 the vertical shear required to produce baroclinic instability is about 10 kt per 4880 m. Therefore on these occasions it was assumed that the air was cloudy with a lapse rate equal to the saturated adiabatic, hence the static stability was taken as zero and then Fleagle's criterion for baroclinic instability was satisfied easily. In those wave depressions with potentially unstable air in their warm sectors lifting makes the term $(1/T)(\mathrm{d}T/\mathrm{d}z + \beta)$ negative, and thus small values of vertical wind shear could lead to rapid deepening. Wave depressions which contain air of low humidity in their warm sectors cannot deepen very much, if at all, because the cooling due to vertical motion must decrease the vertical wind shear which caused the initial baroclinic instability. Of the two depressions which did not deepen when Fleagle's criterion was satisfied, one contained air having a temperature of −10°C, at 500 mb over the warm sector, a circumstance which, according to George and Wolff,[3] inhibits deepening presumably because air having this high temperature has usually subsided and is therefore comparatively dry; while the other was to the west of a cold trough and moving south-eastwards into a baroclinically stable region. Of course development would not have been expected in this case according to the well-known thickness-pattern theory of Sutcliffe and Forsdyke[4].

In order to give a more complete test of Fleagle's theory it was decided to test the behaviour of some summer depressions. Various examples during July, August and September 1957 and June, July and August 1958 were examined giving another 39 occasions. Of these 39 depressions, 4 deepened when the criterion (expression (2)), using the static stability appropriate to cloudless air, was not satisfied. However, all 4 contained cloudy or unstable air in their warm sectors, and Fleagle's criterion was easily satisfied when the static stability was made zero. Of course it is feasible that on these occasions the static stability of a fairly deep layer in the middle troposphere was significantly less than the mean static stability from near sea level to 500 mb One depression failed to deepen (when Fleagle's criterion was satisfied) possibly because there was dry air in the warm sector, a fact confirmed by a lack of cloud ahead of the warm front. The result of the test of Fleagle's criterion is given in Table I.

TABLE I—TEST OF FLEAGLE'S CRITERION FOR BAROCLINIC INSTABILITY

1	2	3
Total number of depressions examined	Number satisfying Fleagle's criterion for deepening	Number not satisfying Fleagle's criterion for deepening
106	101	5
	(a) 96 deepened	*(a)* 5 did not deepen
	(b) 5 did not deepen	*(b)* 0 deepened

Note : Of the 96 depressions examined in *(a)* of column 2, 91 deepened appreciably. In 21 of the 96 the criterion was not satisfied for cloudless air but was easily satisfied for cloudy air.

In *(b)* of column 2 the occasions which did not deepen when the criterion for deepening was satisfied. The lack of deepening was due to very warm and probably very dry air in their warm sectors. One depression was on the western side of the 500 mb trough, a position unfavourable for deepening. There were no occasions, as shown by *(b)* of column 3, of wave depressions deepening when the criterion for baroclinic instability was not satisfied.

Excess vertical wind shear and rate of deepening. When the data for the depressions considered in this note were extracted, the actual deepening which occurred in 24 hours (this deepening could only have been accurate to about 2 mb) was noted and, in order to allow a forecast of deepening in 24 hours to be made, the results have been presented in graphical form in Figure 4 in which excess vertical wind shear (i.e. the wind shear in 4880 m minus the critical shear from Figure 2) has been taken as abscissa and the deepening in 24 hours as ordinate. It is clear that a relationship exists between excess vertical wind shear and the deepening in 24 hours even though the points on the graph are rather scattered. This scattering is not surprising in view of some of the uncertainties involved; for example, the static stability in the middle troposphere in some instances may be significantly different from the mean static stability from near sea level to 500 mb; also during the 24 hours the static stability can drop to zero due to the formation of cloud and thus the excess shear can change considerably during the 24 hours of a forecast period.

In Figure 4, line XA delineates extreme deepening for unstable warm sectors and the line XB gives corresponding values for average deepening. The equations of the line are,

$$\text{for XA} \qquad \Delta p = \frac{4}{3} S$$

$$\text{for XB} \qquad \Delta p = \frac{4}{7} S \qquad \ldots\ldots\ldots\ldots (4)$$

in which S is the excess shear in knots per 4880 m (16 000 ft) and Δp the deepening in 24 hours in millibars.

It may be of interest to note that the largest deepening which occurred in 24 hours in the winter months considered was 40 mb with an excess shear of 46 kt; while the greatest deepening in the summer months was 25 mb in conjunction with an excess shear of 52 kt. The largest excess shear in the winter months was 65 kt and 52 kt in 4880 m in the summer months, but apart from this one value, which occurred in September 1957, all the others during the summer were less than 40 kt.

The effect of horizontal cross-wind shear on baroclinic instability. Fleagle's original criterion contains a term involving the latitudinal shear of the zonal current U given by $dU/d\varphi = -aK\sin\varphi$ and forms part of the inequality (expression (2)). It can be readily shown using Fleagle's original criterion for baroclinic instability that cross-wind cyclonic shear increases the tendency for baroclinic instability to occur (a fact well known to forecasters), since cyclonic shear decreases the right-hand side of inequality in expression (2). However, Fleagle's theory does not allow for any value of cyclonic shear and since it is difficult to estimate this parameter over oceans, its effects will not be developed further at this time, but it is worth noting that large values of cyclonic cross-wind shear can increase the tendency for baroclinic instability considerably.

Conclusions and remarks.

(i) The foregoing test of Fleagle's criterion for the existence of baroclinic instability gave remarkable agreement with theory. Of the 96 occasions of deepening depressions considered (see note to Table I), nearly 80 per cent of them satisfied Fleagle's criterion appropriate to cloudless air, while on 21 occasions Fleagle's criterion was not satisfied in cloudless air but was easily satisfied for cloudy air, the cloud being confirmed by observations.

(ii) It is believed that a large part of the scatter of the points in Figure 4 could be removed if *(a)* the actual lapse rate from say 700 to 500 mb were used instead of a mean from sea level to 500 mb, and *(b)* when the atmosphere is unstable this instability were allowed for as wind shear.

(iii) It would be advantageous when estimating the deepening of a wave depression to find the excess shear both at the chart time and at the expected position of the depression 24 hours later, as baroclinic instability could develop over a wave as it moved north-eastwards.

(iv) It is suggested that the deepening of depressions could be estimated using *(a)* Figure 4, *(b)* George and Wolff's method, and *(c)* pressure tendencies in the warm sector corrected for cross-isobar movement of the depression using synoptic charts. It would also be of interest to check the results of using methods *(a)*, *(b)* and *(c)* with computer-produced values.

(v) It is a matter of experience that summer depressions in temperate latitudes do not, on the average, deepen as quickly or as much as winter ones. This fact is apparent from Figure 4 and is deducible from Fleagle's work since, on the average, summer baroclinic zones are less intense and narrower than winter ones. George and Wolff's method of estimating the deepening of depressions over the North Atlantic Ocean allows for these factors.

Acknowledgement. The writer wishes to thank Mr P. Collison for obtaining the results given in Figure 2 using the Meteorological Office computer.

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NOTES AND NEWS

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Snow cylinders at Geltwil, Switzerland

Plates III and IV represent a very uncommon natural phenomenon originating from meteorological conditions: snow cylinders on sloping terrain, formed presumably under the influence of wind-drag.

The author of the photographs, Mr Arthur Brühlmeier, teacher at Geltwil (Ct. Argovia, Switzerland), was kind enough to communicate his observation to the Schweizerische Meteorologische Zentralanstalt, who have submitted them for publication together with a short meteorological comment.

The chief meteorologist of the Swiss Federal Institute of Snow and Avalanche Research says that in 40 years he has observed only once a similar phenomenon on the Swiss Plateau, and in a less-developed state. Thus the phenomenon in itself may be considered as exceptional.

Date of observation : 5 January 1968, a.m.

Location : Geltwil, co-ordinates 47°15′N, 8°20′E, 625 m, a country village 20 km north of Lucerne, situated on the eastern slope of a hill (Lindenberg, 820 m, running SSE–NNW). Heights are above MSL.

In the immediate vicinity of the point of observation, the slope of the terrain is approximately 1:6, exposure to ENE, meadowland with some fruit trees.

Meteorological observations. The nearest climatological station is Muri 47°16′N, 8°20′E, 540 m, 2·4 km north of Geltwil. The following meteorological data were reported at Muri :

Date	Time	Temperature Dry-bulb	Min.	Max.	Wind Beaufort Force	Relative humidity	Precipitation	Snow depth
	GMT	*degrees C*				*per cent*	*mm*	*cm*
4 Jan.	0630	+ 1·8	− 3·4		W 3–4	76		16
	1230	+ 0·8			NW 5–6	88		
	2030	+ 0·9	− 0·4	+ 2·5	W 1–2	77		
5 Jan.	0630	− 0·8	− 1·0		SE 1–2	92	1·2	22
	1230	+ 3·8			NW 5–6	86		

A good approximation of the temperatures occuring at Geltwil can probably be obtained by subtracting 0·6 degC from the temperatures measured at Muri.

At Zurich Meteorological Institute, strong westerly winds were recorded on 5 January from 00 to 20 hours; they reached 17 m/s at moments.

Synoptic situation. A cold front bringing polar maritime air passed the region from WNW on 4 January before noon. It was accompanied and followed by some wet snow falling on a pre-existent, probably harsh, snow cover.

Characteristics of the snow cylinders. Diameter approximately 50 cm. Note hollow in the centre !

Probable causes of formation. Any human interference must be excluded. Westerly gusts seem to have detached patches of wet snow from the underlying harsh snow surface and rolled them down the slope.

Zurich, Switzerland W. KUHN

REVIEW

Atmospheric electricity, second edition, by J. Alan Chalmers. 25·5 cm × 15 cm pp. x+515, *illus.*, Pergamon Press Ltd, Headington Hill Hall, Oxford, 1968. Price: £6.

This is the second edition of a book first published in 1956 and has been brought up to date by the inclusion of the results of published research up to August 1965. As with the first edition it is rather disappointing to find that 'atmospheric electricity' is defined as 'electrical phenomena between the electrosphere and the earth.' The electrosphere is defined as the first layer above the ground in which the electrical conductivity is high enough for the layer to be effectively a conducting layer, or layer of equipotential. This occurs at a height of about 50 km and below the ionosphere so that all discussion of the ionosphere and electrical effects in the high atmosphere such as aurora are excluded. However, this convention even though disappointing appears to be widely accepted (see *Meteorological glossary*) and so the book cannot really be criticized on this account.

After a preliminary discussion of the history of the subject, general principles and the relevant cloud physics, there are chapters ion the vertcal potential gradient — almost entirely at the earth's surface — space charge, conductivity of the air and the currents which flow between earth and atmosphere. Cloud electricity and the lightning discharge are then discussed, and the chapter on theories of the separation of charge is comprehensive and up to date without, not surprisingly, coming down firmly in favour of any particular theory. The acceptance of the certain existence of 'warm-cloud' thunderstorms came as something of a surprise.

The book is essentially a reference book and virtually every scientific paper on the subject finds mention, more than a thousand of them. Inevitably in parts the text reads more like a catalogue, and one is impressed more by the number of people who have attempted the same thing than by the results of their efforts. Indeed the over-riding impression left by the book is one of much effort about little.

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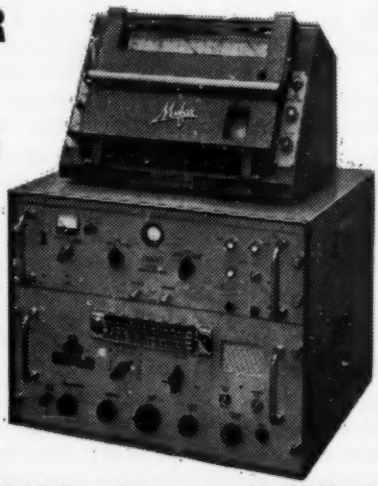

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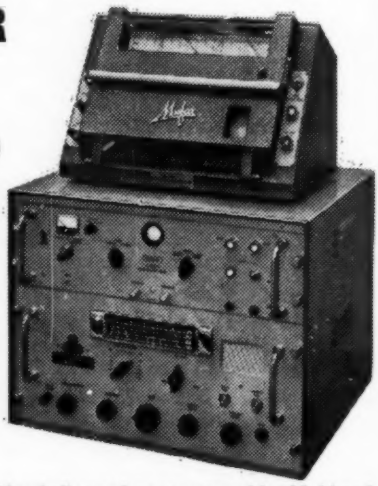

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